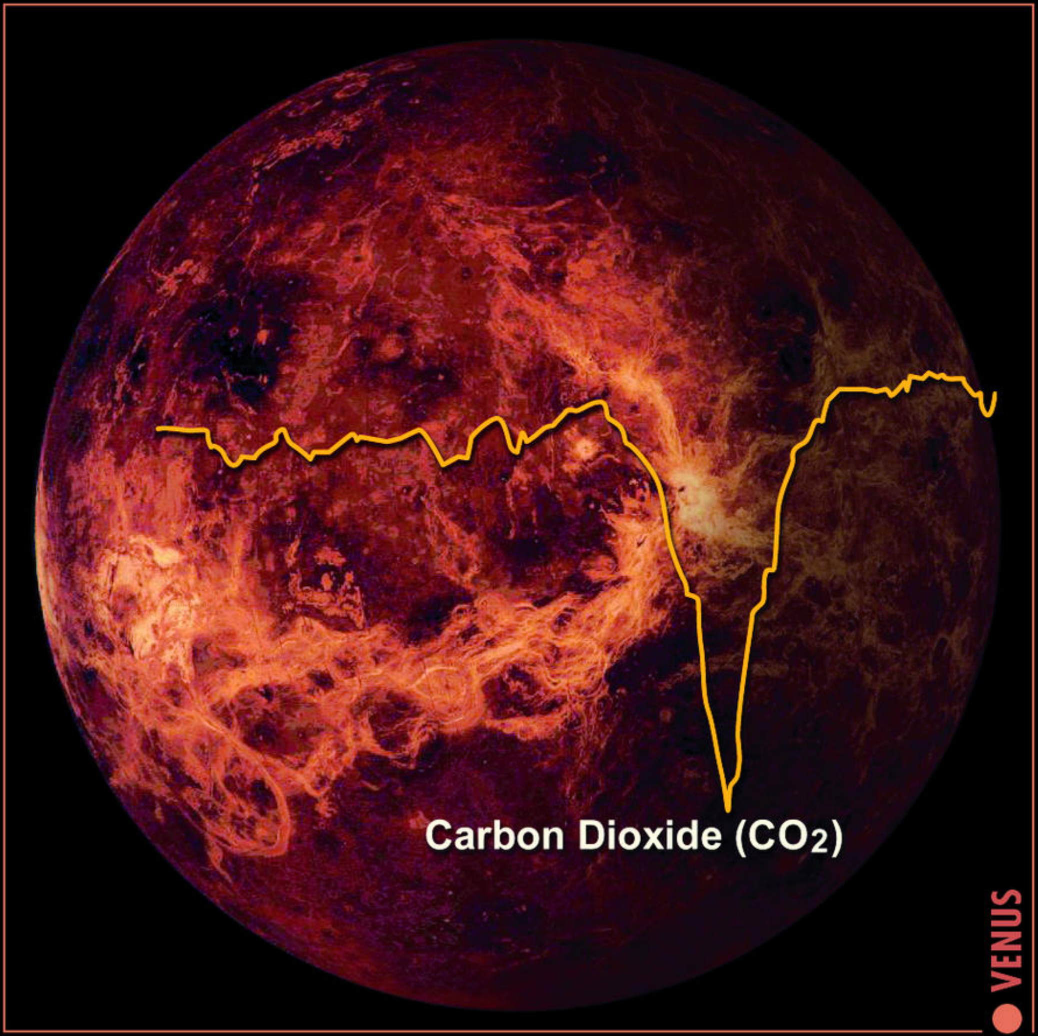
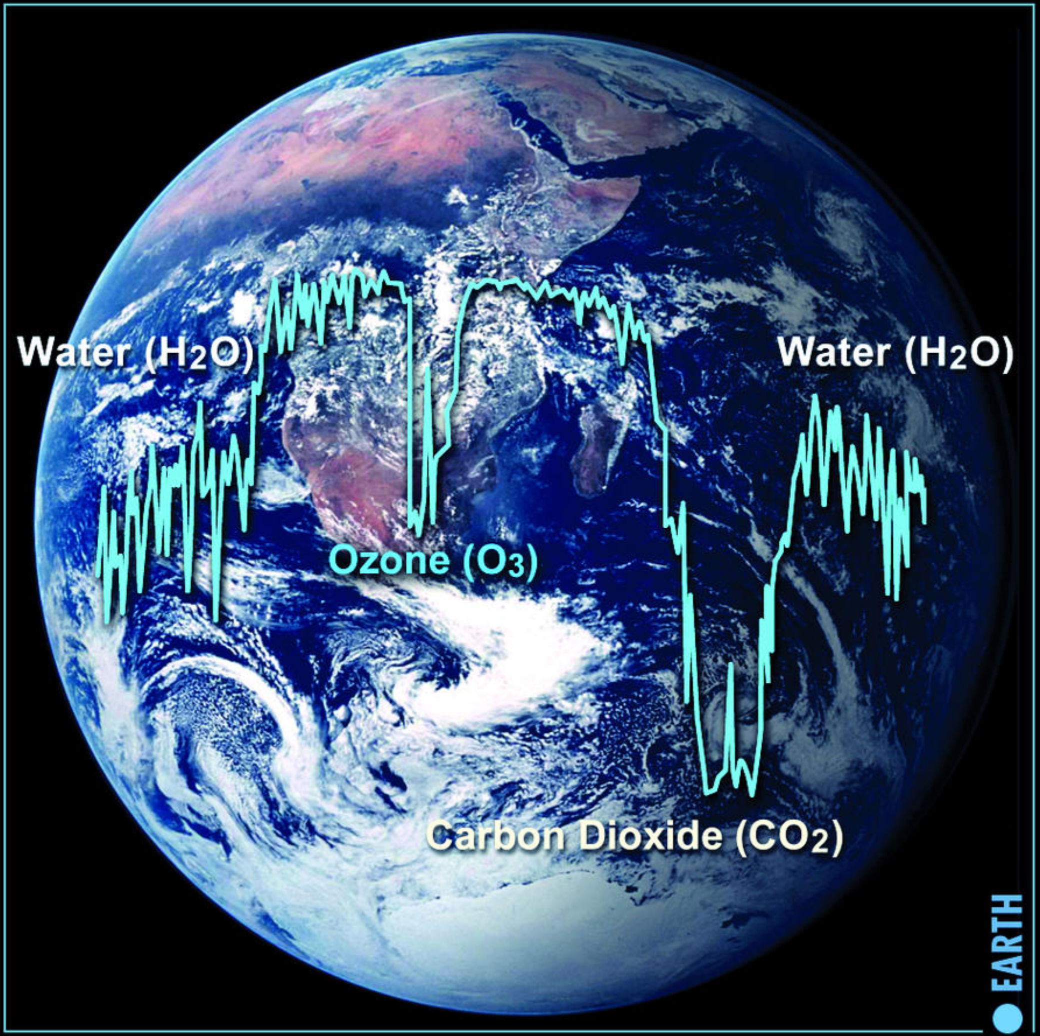
Planetary Habitability





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Topics

HOW TO CHARACTERIZE THE ATMOSPHERE OF A TRANSITING EXOPLANET

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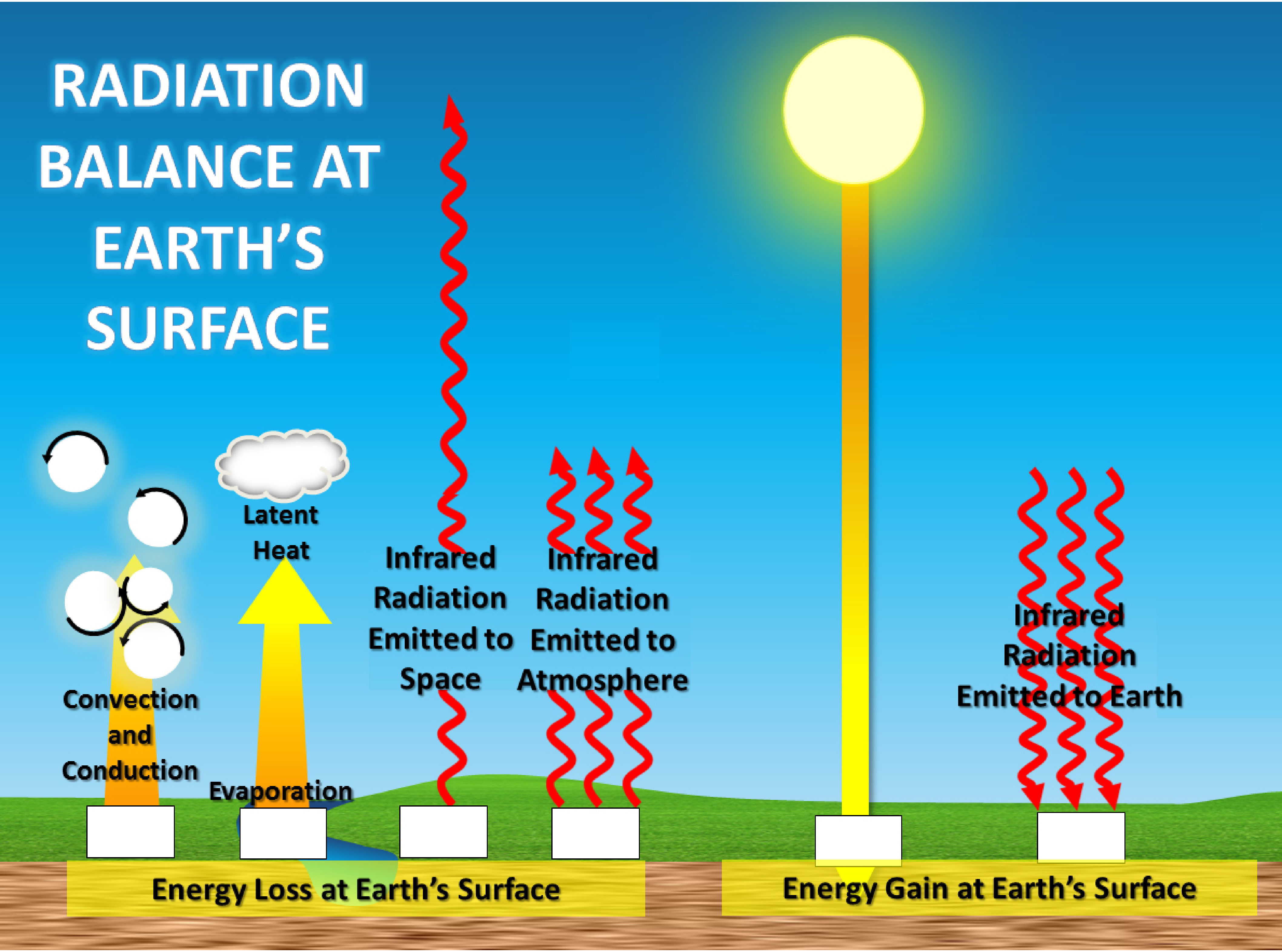
ABSTRACT

This tutorial is an introduction to techniques used to characterize the atmospheres of transiting exoplanets. We intend it to be a useful guide for the undergraduate, graduate student, or postdoctoral scholar who wants to begin research in this field, but who has no prior experience with transiting exoplanets. We begin with a discussion of the properties of exoplanetary systems that allow us to measure exoplanetary spectra, and the principles that underlie transit techniques. Subsequently, we discuss the most favorable wavelengths for observing, and explain the specific techniques of secondary eclipses and eclipse mapping, phase curves, transit spectroscopy, and convolution with spectral templates. Our discussion includes factors that affect the data acquisition, and also a separate discussion of how the results are interpreted. Other important topics that we cover include statistical methods to characterize atmospheres such as stacking, and the effects of stellar activity. We conclude by projecting the future utility of large-aperture observatories such as the James Webb Space Telescope and the forthcoming generation of extremely large ground-based telescopes.

Keywords:


RADIATION BALANCE AT EARTH'S SURFACE
Latent Heat
Convection and Conduction
Evaporation
Infrared Radiation Emitted to Space
Infrared Radiation Emitted to Atmosphere
Infrared Radiation Emitted to Earth
Energy Loss at Earth's Surface
Energy Gain at Earth's Surface

Effects of Obliquity and Eccentricity

The stellar flux received at a particular location is given by

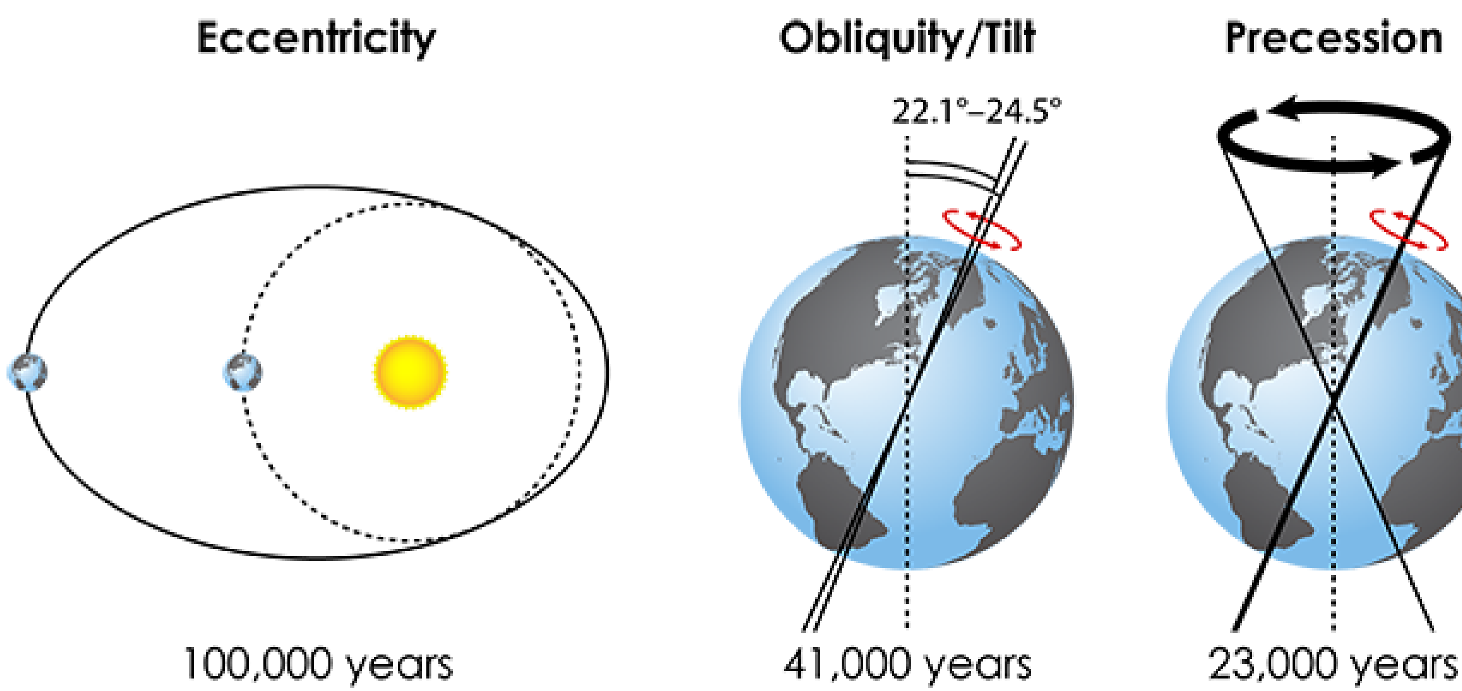
$$F = \frac{L_*}{4\pi a^2} \qquad L_\odot = 3.828\times10^{26}\,W \qquad F_\oplus = 1366\,W m^{-2}$$

more generally written as

$$F = \frac{L_*}{4\pi r^2} \qquad r = \frac{a(1-e^2)}{1+e\cos f}$$

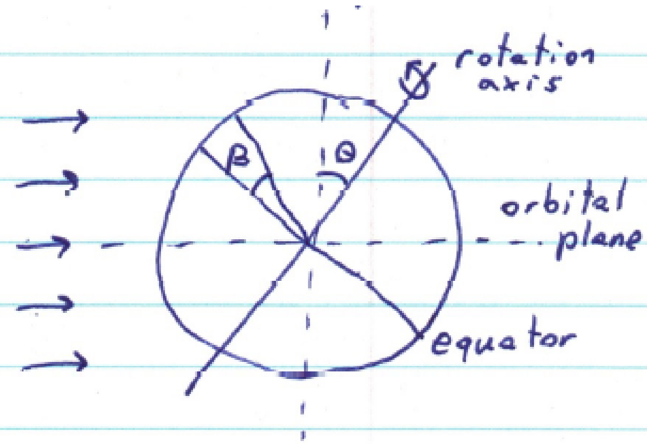
r = star-planet separation
f = true anomaly

Recall: $r = a(1-e)$ at periastron and $r = a(1+e)$ at apastron.



θ = obliquity (angle between rotation axis and normal to orbital plane).

β = latitude



The solar declination is the latitude of the subsolar point and is given by

$$\delta = \theta \cos\left[2\pi(\phi - \Delta\phi)\right]$$

where ϕ is the orbital phase and $\Delta\phi$ is the offset in phase between periastron and highest solar declination in the northern hemisphere. For the Earth, $\theta = 23.5^\circ$ and $\Delta\phi = 0.46$.

The maximum flux at a given latitude is then

$$F = \frac{L_*}{4\pi r^2}(\sin\delta \sin\beta + \cos\delta \cos\beta)$$

$$= \frac{L_*}{4\pi r^2} \cos|\beta - \delta|$$

To include diurnal effects, we need to consider the hour angle of the star with respect to the local meridian, h.

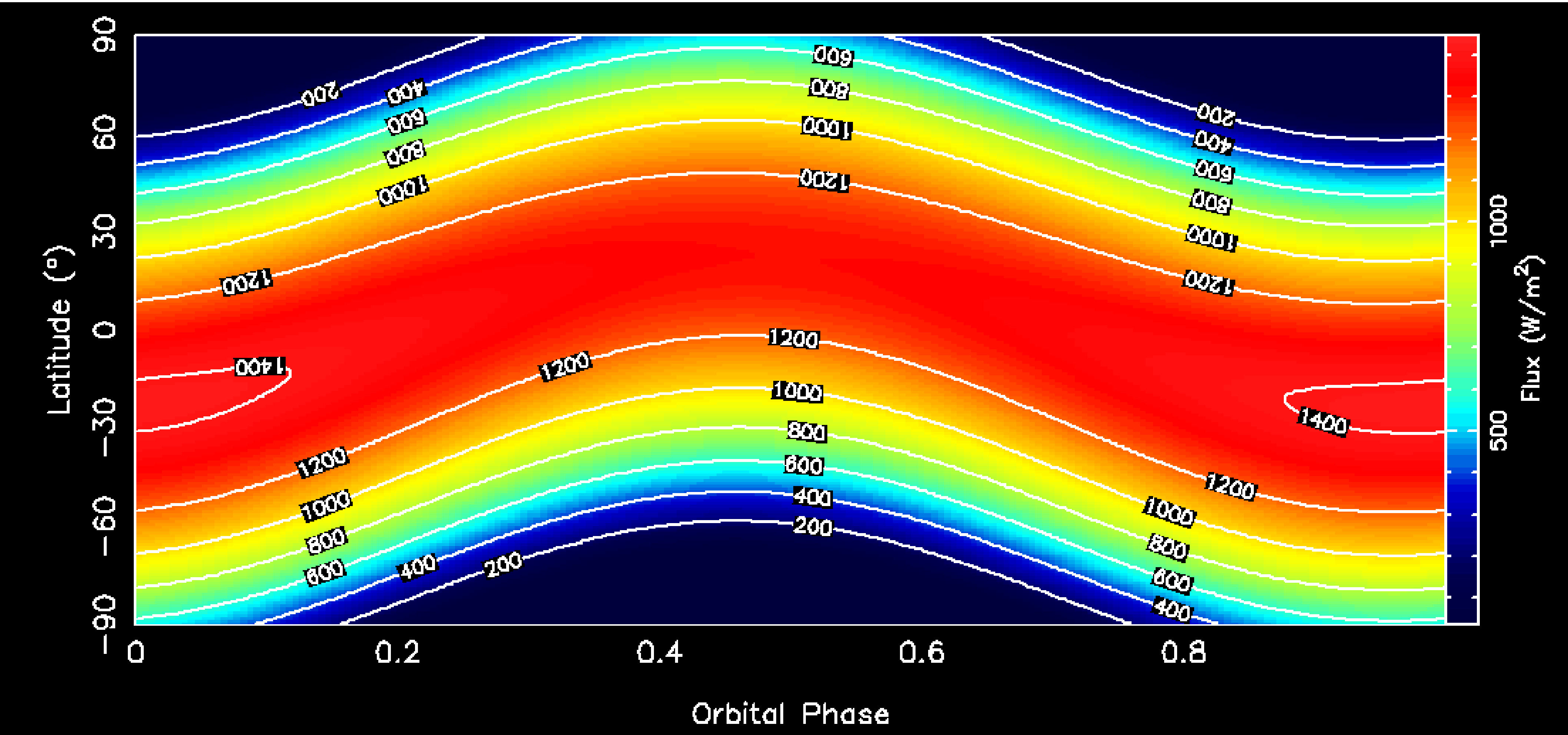
$$F = \frac{L_*}{4\pi r^2}(\sin\delta \sin\beta + \cos\delta \cos\beta \cos h)$$

Also account for the fraction of planetary rotation period that experiences daylight at a given latitude:

$$\Delta t_{dl} = \frac{2 \arccos(-\tan\delta \tan\beta)}{360^\circ}$$

If $\beta + \delta > 90^\circ$ or $\beta + \delta < -90^\circ$ then $\Delta t_{dl} = 1.0$ (constant day)

If $\beta - \delta > 90$ or $\beta - \delta < -90^\circ$ then $\Delta t_{dl} = 0.0$ (constant night)



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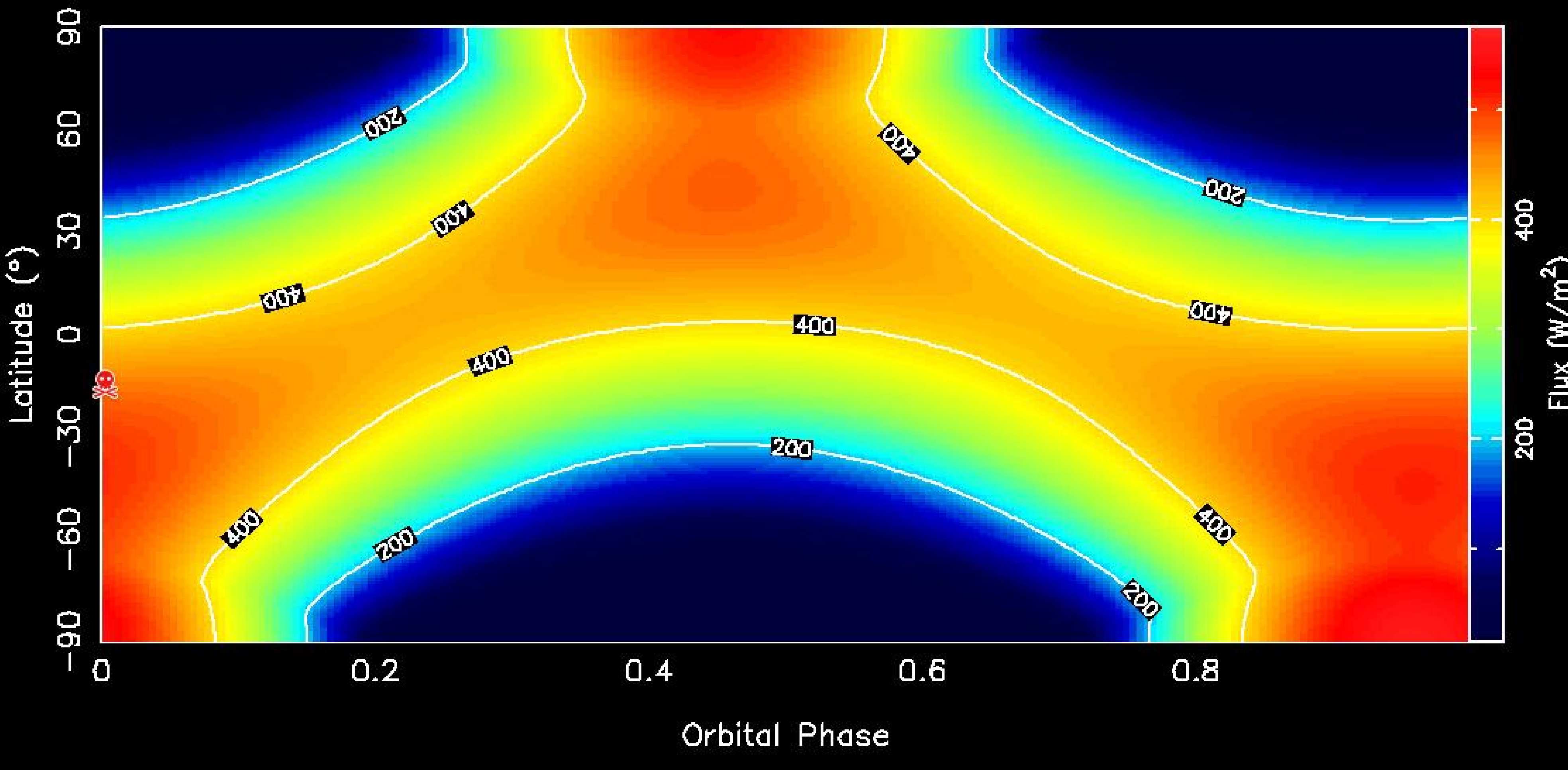
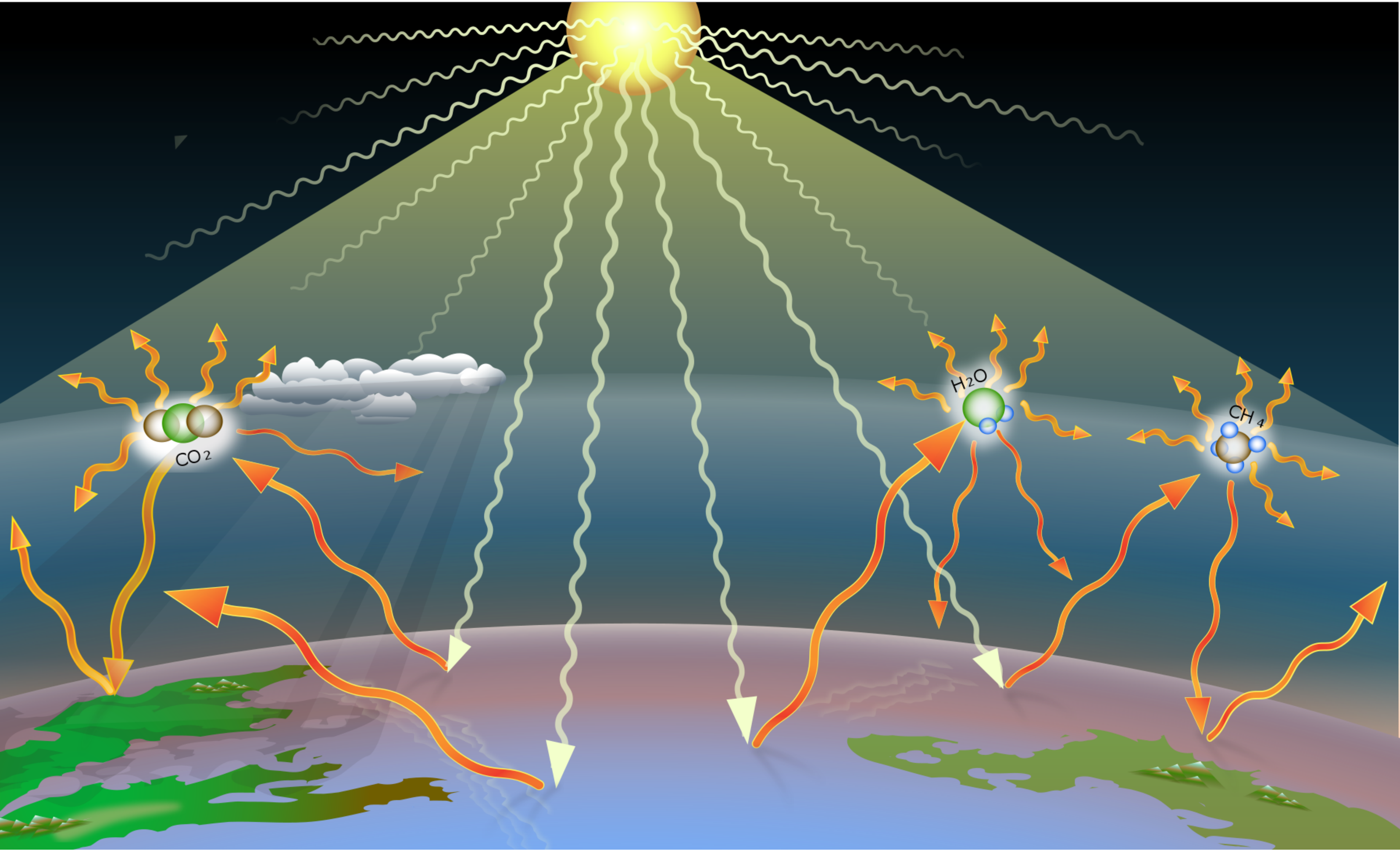




CO_2
H_2O
CH_4

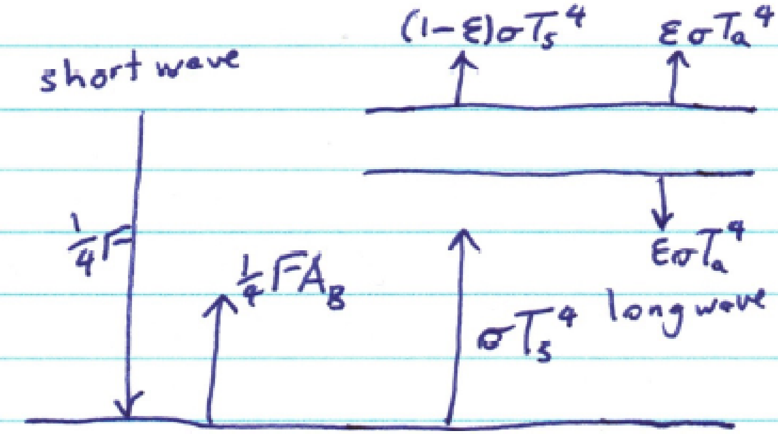
Idealized Greenhouse Model

Recall the planetary equilibrium temperature:

$$T_p = \left[\frac{F(1-A_B)}{4\sigma}\right]^{1/4}$$

For Earth, this results in $T_p = 255\,K = -18\,C$ because it does not account for greenhouse warming.

Emissivity, ε, is the ratio of the energy radiated to that radiated by a blackbody at the same temperature and wavelength.



$$\frac{1}{4}F(1-A_B) = \sigma T_p^4$$

T_s = temp. surface

T_a = temp. atmosphere

IR flux from top of atmosphere is

$$F_\uparrow = \varepsilon\sigma T_a^4 + (1-\varepsilon)\sigma T_s^4$$

Zero net radiation leaving the top of the atmosphere requires

$$-\frac{1}{4}F(1-A_B) + \varepsilon\sigma T_a^4 + (1-\varepsilon)\sigma T_s^4 = 0$$

Zero net radiation entering the surface requires

$$\frac{1}{4}F(1-A_B) + \varepsilon\sigma T_a^4 - \sigma T_s^4 = 0$$

Combining these equations gives

$$2\varepsilon\sigma T_a^4 - \varepsilon\sigma T_s^4 = 0$$

$$\Rightarrow \quad T_a = \frac{T_s}{2^{1/4}}$$

Thus

$$\frac{1}{4}F(1-A_B) = \left(1-\frac{\varepsilon}{2}\right)\sigma T_s^4$$

$$T_s = \left[\frac{F(1-A_B)}{4\sigma}\,\frac{1}{1-\frac{\varepsilon}{2}}\right]^{1/4}$$

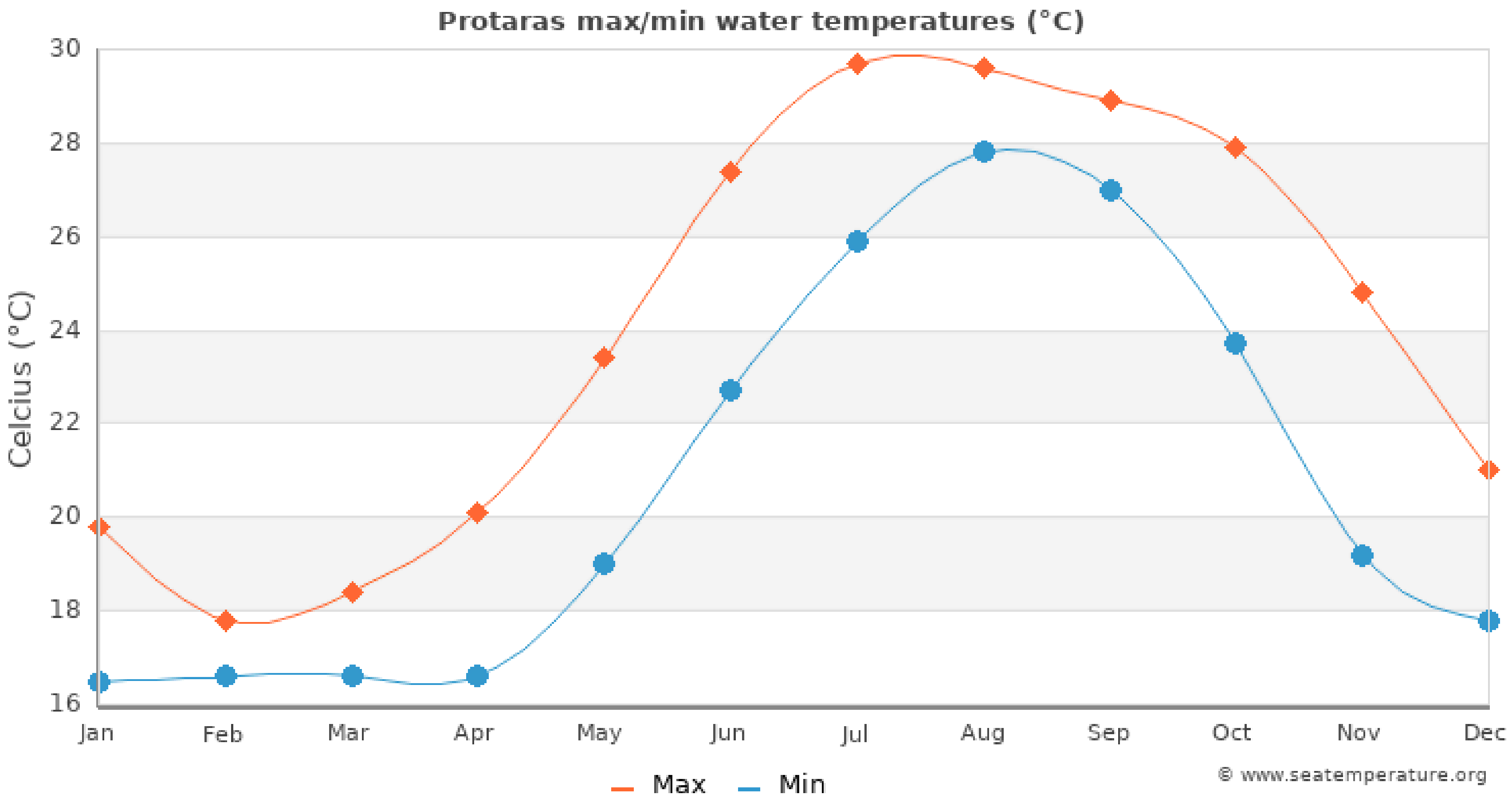
For $\varepsilon = 0$, $T_s = 255K$ (same as before)
$\varepsilon = 1$, $T_s = 303k = 30\ C$
$\varepsilon = 0.78$, $T_s = 288K$, $T_a = 242K$
] For Earth

Assumptions:

1. Assumes the atmosphere is isothermal. The troposphere is the most important greenhouse layer in the atmosphere, where temperature decreases with height. Thus more radiation is emitted downward then upward.

2. Assumes the atmosphere absorbs all outgoing radiation at all IR wavelengths, whereas the absorption is highly wavelength dependent.

3. Conduction of heat and evaporation of water tranfer twice as much energy to the atmosphere than radiative transfer.


Protaras max/min water temperatures (°C)
Celcius (°C)
30
28
26
24
22
20
18
16
Jan
Feb
Mar
Apr
May
Jun
Jul
Aug
Sep
Oct
Nov
Dec
Max
Min
© www.seatemperature.org

Radiative Equilibrium Timescale

From above, energy balance can be written as

$$\frac{1}{4}F(1-A_B) + \varepsilon\sigma T_a^4 = \sigma T_s^4 \qquad \text{— (1)}$$

For an atmospheric column of unit cross-sectional area in hydrostatic balance, the time dependent energy balance is

$$\frac{c_p P_s}{g}\frac{dT_a}{dt} = \varepsilon\sigma T_s^4 - 2\varepsilon\sigma T_a^4 \qquad \text{— (2)}$$

where c_p is the heat capacity at constant pressure, P_s is the pressure at the surface, and g is surface gravity.

The mass of the atmospheric column per unit area is P_s/g. Assume the atmospheric temperature can be expressed as

$$T_a = T_{a0} + \Delta T$$

where $|\Delta T| \ll T_{a0}$ and T_{a0} is the radiative equilbrium temperature of the atmosphere.

To determine the time evolution of the temperature perturbation, ΔT, substitute equation (1) into (2):

$$\frac{c_p P_s}{g}\frac{d\Delta T}{dt} = \varepsilon(1-A_B)\frac{F}{4} + (\varepsilon-2)\varepsilon\sigma(T_{a0}+\Delta T)^4 \qquad \text{— (3)}$$

Using a Taylor series truncated after the second term:

$$f(T_{a0} + \Delta T) = f(T_{a0}) + \Delta T \frac{df(T_{a0})}{dT}$$

then equation (3) becomes:

$$\frac{c_p P_s}{g}\frac{d\Delta T}{dt} = \varepsilon(1-A_B)\frac{F}{4} - (2-\varepsilon)\varepsilon\sigma T_{a0}^4 - 4(2-\varepsilon)\varepsilon\sigma T_{a0}^3 \Delta T$$

$$= -\varepsilon^2\sigma T_{a0}^4 + \varepsilon\sigma T_s^4 - (2-\varepsilon)\varepsilon\sigma T_{a0}^4 - 4(2-\varepsilon)\varepsilon\sigma T_{a0}^3 \Delta T$$

$$= -\varepsilon^2\sigma T_{a0}^4 + 2\varepsilon\sigma T_{a0}^4 - 2\varepsilon\sigma T_{a0}^4 + \varepsilon^2\sigma T_{a0}^4 - 4(2-\varepsilon)\varepsilon\sigma T_{a0}^3 \Delta T$$

Thus

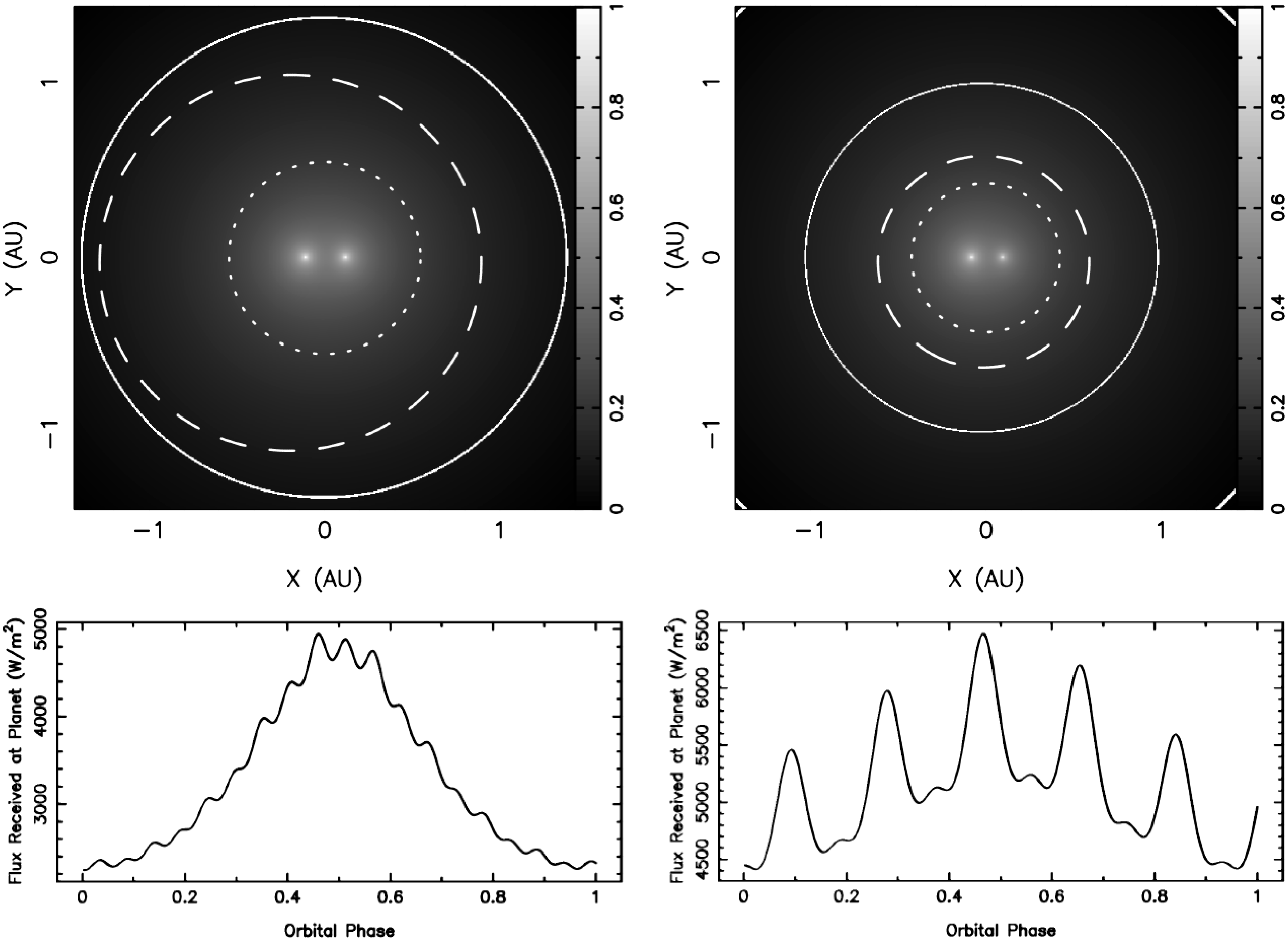
$$\frac{d\Delta T}{dt} = -\frac{4(2-\varepsilon)\varepsilon\sigma g T_{a0}^3}{c_p P_s}\Delta T$$

Integrating both sides yields the radiative equilibrium timescale:

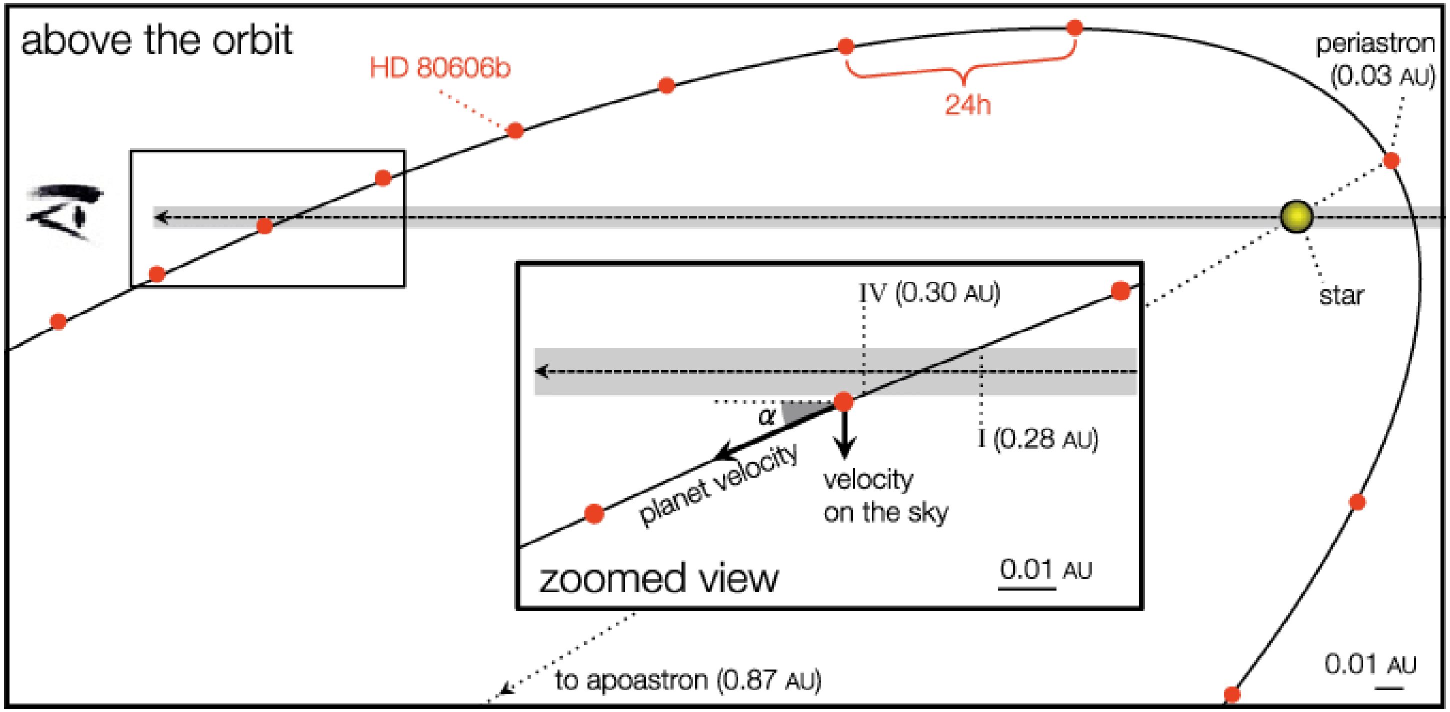
$$\tau_E = \frac{P_s c_p}{4(2-\varepsilon)\varepsilon\sigma g T_{a0}^3}$$

This represents the time required for the atmosphere to respond significantly to changes is radiative forcing and is an expression of the thermal inertia of the atmosphere.

For $T_{a0} = 300K$, $\varepsilon = 0.7$, $g = 10\,m s^{-2}$, $P_s = 10^5\,Pa$, $c_p = 1000\,J K^{-1} kg^{-1}$; we find $\tau_E = 20$ days.



Y (AU)
X (AU)
Y (AU)
X (AU)
Flux Received at Planet (W/m^2)
Orbital Phase
Flux Received at Planet (W/m^2)
Orbital Phase


above the orbit
HD 80606b
24h
periastron
(0.03 AU)
star
IV (0.30 AU)
I (0.28 AU)
α
planet velocity
velocity
on the sky
zoomed view
0.01 AU
to apoastron (0.87 AU)
0.01 AU